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Mass/Age Distribution and Composition of Sediments on the Ocean Floor and the Global Rate of Sediment Subduction

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The total mass of sediments on the ocean floor is estimated to be 262×10^{21} g. The overall mass/age distribution is approximated by an exponential decay curve: $(11.02 \times 10^{21}\ \mathrm{g})e^{-0.0355t}$ Ma. The mass/age distribution is a function of the area/age distribution of ocean crust, the supply of sediment to the deep sea, and submarine erosion and redeposition. About 140×10^{21} g of the sediment on the ocean floor is pelagic sediment, consisting of about 74% $CaCO_3$, with the remainder opaline silica and red clay. Of the sediment on the ocean floor, 122×10^{21} g is detritus, mostly terrigenous, but a small portion (about 6×10^{21} g) is volcanic. Because very little pelagic sediment is obducted, virtually all of the pelagic sediment mass and some fraction of the terrigenous sediment is being subducted at a rate estimated to be about 1×10^{21} g per million years. The composition of sediment on the ocean floor differs significantly from that of average passive margin and continental sediment, so that the loss of ocean floor sediment through subduction may drive the composition of global sediment toward enrichment in silica, alumina, and potash and toward depletion in calcium.

INTRODUCTION

The terms "ocean floor sediments" and "pelagic sediments" are often used interchangeably, as though they were synonymous. In discussing the works of previous authors, we use their terminology in referring to either ocean floor sediments or pelagic sediments. The term "pelagic" sediment was coined by *Murray and Renard* [1884, 1891] to designate those sediments which accumulate on the ocean floor by settling through the water column. Originally, it was assumed that the particles settled separately through the water, but more recently it has become evident that most particles settle in aggregates, usually as fecal pellets [*Honjo*, 1977; *Brewer and Glover*, 1987]. In addition to the pelagic sediment proper, the ocean floor receives "terrigenous" sediments [*Murray and Renard*, 1884, 1891] from land or island sources, which arrive directly at the bottom of the ocean without settling through the water column. These are carried downslope and to the deep seafloor horizontally by turbidity currents [*Hay*, 1974], but even upslope bottom current transport processes operate, forming sediment drifts [*Kidd and Hill*, 1986]. The term "ocean floor sediments" has generally excluded the sediments of the continental margin inshore of the base of the continental rise. Recently, as a result of analysis of Deep Sea Drilling Project (DSDP) data [*Sloan*, 1985] and extensive seismic surveying, information has become adequate to enable us to estimate the relative contributions of these two sedimentation mechanisms to the total ocean floor sediment mass.

PREVIOUS ESTIMATES

Volume or Mass of Sediments on Ocean Floor

The previous estimates of quantities of ocean floor sediments can be divided into two categories: (1) estimates of total bulk volume or mass and (2) estimates which attempted to take the mass/age distribution into account.

There have been four modern estimates of the total volume or mass of "pelagic sediment" or "sediment on the ocean floor." *Poldervaart* [1955] estimated the area covered by pelagic sediments to be 268×10^6 km^2 and their average thickness to be 0.6 km, including an average 50% pore space, for a total column of 80.4×10^6 km^3 solid or a mass of 217×10^{21} g. *Ronov and Yaroshevsky* [1969, 1977] estimated the area of pelagic sedimentation to be 300×10^6 km^2 and the average thickness to be 0.4 km, and they assumed an average porosity of 40%, for a total volume of 72×10^6 km^3 or a mass of 190×10^{21} g. *Southam and Hay* [1977] estimated the area over which pelagic sediments settle to be 311×10^6 km^2 and assumed an average thickness of 0.5 km and an average porosity of 25%, which indicates a volume of 116.6×10^3 km^3 solid; they cited a mass of 356×10^{21} g. *Southam and Hay* [1981] used the same area and thickness, but they used 35% for the average porosity, estimating a volume of 92×10^6 km^3 solid and a mass of 249×10^{21} g.

The most accurate estimate of the total volume of sediment on the ocean floor has recently been presented by *Howell and Murray* [1986], who measured the isopach maps of *Emery and Uchupi* [1984] for the Atlantic Ocean, including the Gulf of Mexico and the Caribbean, of *Tucholke et al.* [1982] for the Northwest Atlantic, of *Ludwig and Houtz* [1979] for the Pacific; and of *Udintsev et al.* [1975] for the Indian Ocean. Their estimate does not include sediment shoreward of the base of the continental rise, and it excludes the Arctic Ocean and the region adjacent to Antarctica. They arrive at a total bulk volume of 141.7×10^6 km^3, with an average porosity of 34% and a solid volume of 94.1×10^6 km^3. Using the results of analyses of DSDP cores [*Miskell et al.*, 1985; *Sloan*, 1985], they estimated the solid phase to be 70.3×10^6 km^3 terrigenous material, 2.9×10^6 km^3 volcanic material (density 2.2 g cm^{-3}), 15.9×10^6 km^3 biogenic $CaCO_3$ (density 2.72 g cm^{-3}), and 5.0×10^6 km^3 biogenic silica (density 2.30 g cm^{-3}). Assuming the density of the terrigenous material to be 2.7 g cm^{-3}, the mass of the solid phase then works out to be 250.9 $\times 10^{21}$ g, excluding the Norwegian-Greenland Sea, the Arctic basin, and a small area of the Atlantic sector of the circum Antarctic.

From the map of *Perry et al.* [1980] and discussion by Perry [1986], we estimate the area of the Norwegian-Greenland Sea seaward of the base of the continental rise to

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be about 1.45×10^6 km^2. The sediment thicknesses in this region are highly variable. [*Gronlie and Talwani*, 1978; *Vogt*, 1986], but an average thickness of 1 km is reasonable. Assuming an average porosity of 35%, this yields an estimate of 0.94×10^6 km^3 solid and, if the solid phase density is 2.7 g cm^{-3}, a mass of 2.54×10^{21} g.

The area of the Arctic basins proper is about 10.07×10^6 km^2. Sediment thicknesses are not well known; the most extensive attempt at sediment isopachs in the Arctic region is on the tectonic map of *Gorshkov* [1980, pp. 26–27]. Although the isopachs are discontinuous or missing from some parts of the Arctic basin, the average sediment thickness is about 1 km. Assuming an average porosity of 35%, this indicates a solid phase volume of 6.55×10^6 km^3 and, assuming a density of 2.7 g cm^{-3}, this implies a mass of 17.67×10^{21} g.

Sediment thickness maps of the Atlantic sector of the Antarctic have recently been prepared [*Rodriguez et al.*, 1986]. The region in this sector south of 60°S, not included in the estimate of *Howell and Murray* [1986], contains sediment having a bulk volume of 2.44×10^6 km^3 and a mass of 4.28 $\times 10^{21}$ g.

The sum of sediment masses in these areas, Norwegian-Greenland Sea (2.5×10^{21} g), Arctic basin (17.7×10^{21} g), and Atlantic sector of the Antarctic (4.3×10^{21} g), is 24.5×10^{21} g, which, added to *Howell and Murray*'s [1986] estimate for the rest of the Atlantic, the Indian, and the Pacific Ocean basins (250.9×10^{21} g), gives a new global total of 275×10^{21} g of sediment on the ocean floor.

Mass/Age Distribution of Sediments on Ocean Floor

Two estimates based on summations of the mass/age distribution of sediment on the ocean floor have been published recently. *Gregor* [1985, Table 1] estimated the existing mass/age distribution of pelagic sediment by assuming an average mean global accumulation rate of 0.6 g cm^{-2} kyr^{-1} and multiplying this by the area of ocean floor still in existence and formed during or before each of 13 unequal time increments, ranging from the present to 180 Ma, based on seafloor magnetic lineations [*Sclater et al.*, 1980]. *Sclater et al.* [1980] had determined the area of surviving seafloor formed during each of these time increments. *Gregor* [1985] determined the total area of surviving seafloor which existed during each of the time increments by adding half of the area of surviving seafloor formed during a given time increment and the entire area formed during each of the preceding time increments back to 180 Ma. The result was an estimate of the total pelagic sedimentary mass of 104.6×10^{21} g. Because the accumulation rate was assumed to be constant, the only remaining factor controlling Gregor's pelagic sediment mass/age distribution was the subduction of ocean floor through time. The assumed mean global accumulation rate of 0.6 g cm^{-2} kyr^{-1} was attributed to *Lisitzin* [1972] and *Worsley and Davies* [1979], and was supposed to be a value for the past 50 m.y. Unfortunately, this assumption of a constant accumulation rate was a great oversimplification, which obscured significant trends in the mass/age distribution and resulted in underestimation of the total mass of existing ocean floor sediment. Subtracting the estimated thickness of the pelagic sediments from 16 seismic profiles across abyssal plains, Gregor estimated the nonpelagic sediments of the abyssal plains to have a mass of 73×10^{21} g (Tertiary) and 39×10^{21} g (Cretaceous), for a total of 112×10^{21} g. Finally, Gregor estimated the volume of deep-sea fans to be 7×10^6 km^3 and their solid phase to be 3.2×10^{21} g (Tertiary) and 2.2×10^{21} g (Cretaceous) for a total of 15.4×10^{21} g. Summing masses for the pelagic sediment and nonpelagic sediments of the abyssal plains and deep-sea fans, Gregor's data indicated a total of 232×10^{21} g of sediment on the ocean floor.

Veizer and Jansen [1985] used a total of 375.9×10^{21} g of "oceanic sediment" (1.1×10^{21} g Jurassic, 90.1×10^{21} g Cretaceous, 284.7×10^{21} g Cenozoic), citing *Gregor* [1985] as their data source. This apparent discrepancy is because Veizer and Jansen's total includes sediment in marginal seas.

Budyko et al. [1985, 1987] estimated the total sediment volume on the ocean floor (excluding the Pleistocene and the sediments of the continental margins) to be 96×10^6 km^3, and the mass to be 170×10^{21} g. *Ronov et al.* [1986*a*, *b*] gave the pre-Pleistocene mass as 177×10^{21} g. *Ronov et al.* [1986*a*] presented their own age/area distribution for the sea floor, which was similar to that of *Sclater et al.* [1980], but the time increments were subperiods of the Jurassic and Cretaceous and epochs of the Cenozoic. In contrast to Gregor, *Ronov et al.* [1986*a*, pp. 3-4 (translation)] noted that

> the sediment volumes are unevenly distributed over stratigraphic subdivisions in the continents, at the margins and in the oceans, although they are strictly consistent. They vary synchronously in those zones and in accordance with a general law.

Their Figures 3 and 5 showed that the fluctuations of sediment volumes and sedimentation rates on the continents, margins, and in the ocean varied by a factor of 2, even when averaged over long time intervals of 20–30 m.y. Their estimate of the mass/age distribution is based on analysis of the age-/area distribution and varying sedimentation rates known from DSDP cores. They estimated the total volume of ocean sediments (including the Quaternary) to be 114×10^6 km^3 and the mass to be 203×10^{21} g, distributed over an area of 280×10^6 km^2. *Budyko et al.* [1985, 1987] presented data for mass/age distribution of sediments on the floors of the oceans in tabular form [*Budyko et al.*, 1987, Table 7]. *Ronov et al.* [1986*a*] constructed their Figure 3 from this data.

The recent estimate of a total of 203×10^{21} g of sediments on the ocean floor by *Ronov et al.* [1986*a*] is somewhat lower than the estimates of *Southam and Hay* [1981] (249×10^{21} g), *Howell and Murray* [1986] (251×10^{21} g), and *Gregor* [1985]; (232×10^{21} g) and probably reflects the differences between isopach maps of the ocean basins constructed from Soviet and American data bases.

A NEW ESTIMATE

To prepare a new estimate of the mass/age distribution of oceanic sediment, we transformed the seafloor area/age data of *Sclater et al.* [1980] into area of seafloor in existence at 5-m.y. intervals from 0 to 180 Ma, excluding marginal seas (as had *Gregor* [1985]). The area seafloor in existence at the midpoint of each 5-m.y. interval was multiplied by the average solid-phase sediment accumulation rate for the interval, as shown in Table 1. The resulting mass/age distribution is displayed as a histogram in Figure 1.

Average solid-phase sediment accumulation rates for epochs and subepochs of the Cenozoic were determined by *Sloan* [1985], who used averages of about 16,000 sediment accumulation rates determined from data for legs 1–79 of the Deep Sea Drilling Project to estimate global sediment accumulation

TABLE 1. Mass/Age Distribution of Sediments on Ocean Floor

Interval, Ma	Area of the Present Ocean Floor in Existence at Midpoint of Interval, 10^{16} cm^2	Accumulation of Sediment Still in Existence for Representing Each 5-m.y. Interval, g cm^{-2}	Total Mass of Sediment in Each 5-m.y. Interval, 10^{21} g	Cumulative Total Mass, 10^{21} g
0–5	273.9	18038.	49.41	49.41
5–10	256.6	8920.	22.89	72.30
10–15	241.8	8558.	20.69	92.99
15–20	228.3	7995.	18.25	111.24
20–25	215.7	7566.	16.32	127.56
25–30	204.0	6865.	14.00	141.56
30–35	192.3	5150.	9.90	151.46
35–40	181.5	6748.	12.25	163.71
40–45	171.7	8510.	14.62	178.33
45–50	161.8	8650.	14.00	192.33
50–55	152.0	7906.	12.02	204.35
55–60	141.6	7208.	10.21	214.56
60–65	131.2	4902.	6.43	220.99
65–70	120.8	6340.	7.66	228.65
70–75	110.3	4152.	4.58	233.23
75–80	99.9	3060.	3.06	236.29
80–85	89.4	3446.	3.08	239.37
85–90	79.8	7536.	6.01	245.38
90–95	70.6	4416.	3.12	248.50
95–100	61.7	3447.	2.13	250.63
100–105	53.4	3460.	1.85	252.48
105–110	45.1	3646.	1.64	254.12
110–115	39.7	3470.	1.38	255.50
115–120	34.3	2590.	0.89	256.39
120–125	29.7	3637.	1.08	257.47
125–130	25.5	4015.	1.02	258.49
130–135	21.3	5831.	1.24	259.73
135–140	17.2	3807.	0.65	260.38
140–145	13.0	3807.	0.49	260.87
145–150	8.8	3807.	0.34	261.21
150–155	6.5	3807.	0.25	261.46
155–160	4.4	3807.	0.17	261.63
160–165	3.0	3807.	0.11	261.74
165–170	2.1	3807.	0.08	261.82
170–175	1.3	3807.	0.05	261.87
175–180	0.4	3807.	0.02	261.89

rates. Sloan's estimates of rates for epochs and subepochs were normalized to the 5-m.y. intervals, using the time scale of *Berggren et al.* [1985*a*] and *Berggren et al.* [1985*b*]. For the Late Mesozoic we determined average accumulation rates for the stages of the Cretaceous using the entire DSDP data base. Thicknesses of strata representing Cretaceous stages at DSDP sites were converted to accumulation rates, using the time scale of *Hallam et al.* [1985] and assuming an average porosity of 30% (determined by inspection of DSDP GRAPE logs) and an average grain density of 2.7 g cm^{-3}. Occasionally, highly anomalous thicknesses for stages are reported; if the thickness at a site was greater than three standard deviations larger than the average of the thicknesses at all of the other sites, it was disregarded. For the Jurassic the Deep Sea Drilling Project data are a poor estimate of ocean-wide sediment accumulation rates, because most of the sites are located in atypical areas and the variance in thicknesses is quite large. We assumed that the average sediment accumulation rates on the Jurassic seafloor (135–180 Ma) are the same as the average for the early Cretaceous, between 100 and 135 Ma. In any case, the area of the ocean floor older than 135 Ma is small (about 6% of the present area), so that an inaccurate estimate of the earlier Mesozoic accumulation rates does not appreciably change the estimate of the total sedimentary mass.

The increments of mass for each 5-m.y. age interval were summed to arrive at an estimate of 262×10^{21} g for the total existing mass of sediment; this is intermediate between the original estimate of *Howell and Murray* [1986] and our reestimating adding the sediments of the Norwegian Greenland Sea, Arctic Ocean, and Atlantic sector of the Antarctic. At this time the estimate of the mass of sediment on the ocean floor is probably more accurate than that for any other major sediment body.

The sediment on the ocean floor can be divided into five components: calcium carbonate, opaline silica, red clay, volcanogenic sediment, and bottom-transported terrigeneous sediment. Unfortunately, the mass/age distribution of these components is incompletely known. The calcium carbonate is mostly of biogenic, pelagic origin, being fixed as calcite by planktonic foraminifers, coccolithophores, and, to a much lesser extent, benthic foraminifers. A very small amount of the carbonate which accumulates is fixed as aragonite by pteropods (pelagic gastropods). Some carbonate accumulating in the deep sea is aragonite and high magnesium calcite shed

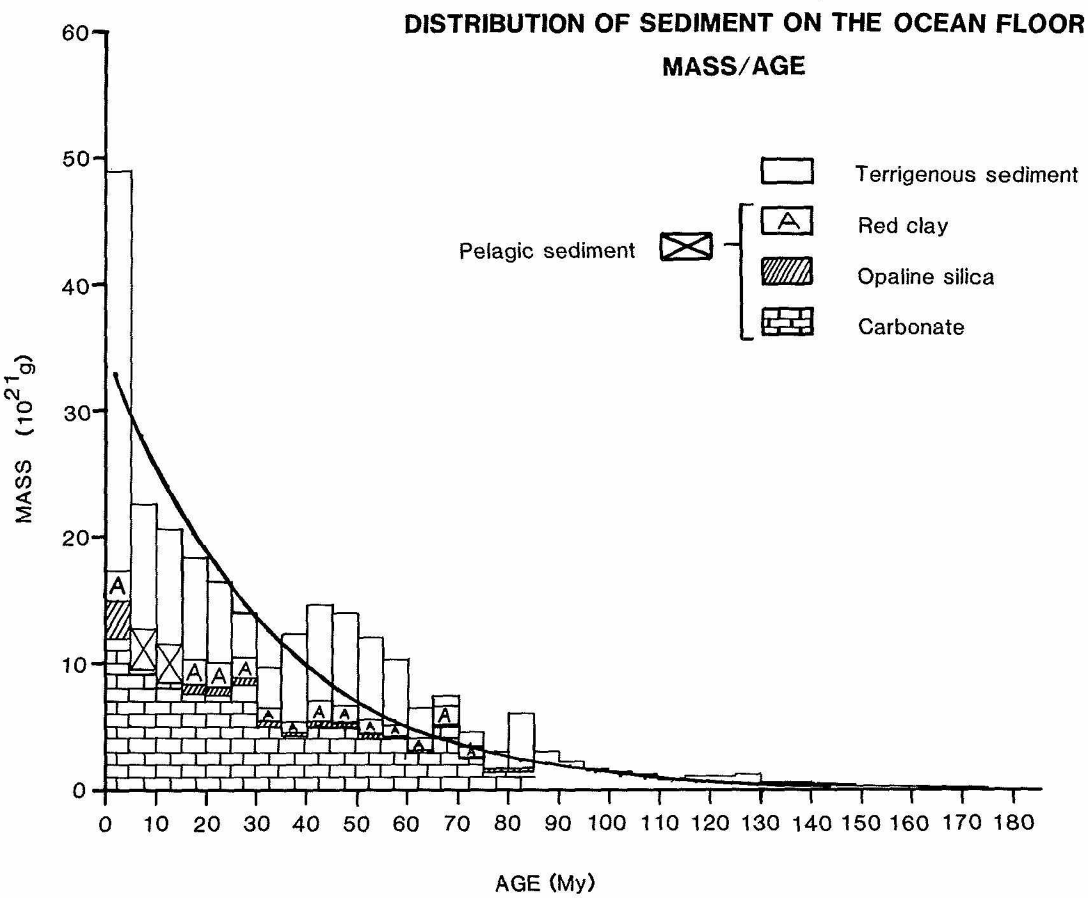

Fig. 1. Mass/age distribution of ocean floor. Terrigenous, red clay, opaline silica, and carbonate components are shown. The solid curve is $(11.02 \times 10^{21}\ \mathrm{g})e^{-0.0355t}$ Ma.

from shallow banks or shelves. Accumulation rates of calcium carbonate for the Cenozoic were determined by *Sloan* [1985]; for the Mesozoic they were interpreted from the sedimentation rate data of *Southam and Hay* [1981].

Opaline silica is almost exclusively of pelagic origin, being fixed by diatoms and radiolarians in the water column, but its abundance is often difficult to determine. *DeMaster* [1981] discussed the flux of opaline silica into and out of the marine environment. Because of the deficiency of data for the last 5 m.y., we have assumed that his estimate of the present-day flux into the ocean (6.1×10^{14} g yr^{-1}) is typical for the last 5 m.y., and that for a steady state, input must equal output. Applying the decay curve which best fits the data, discussed below, the mass of opaline silica representing the last 5 m.y. should be about 2.791×10^{21} g. Masses of opaline silica in ocean floor sediments from 15 to 85 Ma were determined from DSDP data by *Miskell et al.* [1985], but ancient coastal upwelling systems were not included. As *Heath* [1974] noted, a major proportion of the opaline silica is presently extracted in coastal upwelling zones and in the Gulf of California [*Calvert*, 1966]. Because deposition of opaline silica is so strongly dependent on upwelling sites, and because such sites are not well known for the past, the values determined by Miskell et al. may represent only a portion of the total flux out of ocean water.

Red clay, considered by *Murray and Renard* [1891] to be of volcanic origin, has traditionally been considered to be a product of aerial transport and subsequent settling through the water column, but it has been recognized that other transport mechanisms might also be involved. *Clauer and Hoffert* [1985] have found that red clay in two cores from the South Pacific consisted of redeposited authigenic sediment which formed at some other location. Clauer and Hoffert concluded that the term "sedimentation rate" is inappropriate for red clays because they may be entirely authigenic, and they suggested that "the appropriate concept is formation and accumulation of authigenic material by time unit."

Pelagic sediment is usually considered to be a three-component system consisting of carbonate, siliceous ooze, and red clay. *Southam and Hay* [1977] discussed the composition of pelagic sediment before and after the appearance of calcareous plankton. Chemical compositions for these components have been suggested by *Sujkowski* [1952] and by *Poldervaart* [1955], who utilized previously unpublished analyses by William Rubey. Poldervaart estimated the relative volumes of carbonate ooze, siliceous ooze, and red clay to be 71.9, 8.9, and 19.2% of the total sediment volume, respectively. *Southam and Hay* [1981] noted that the average of several thousand samples of pelagic sediment from the Deep Sea Drilling Project was 74% $CaCO_3$ and hence Poldervaart's estimate still stands up very well as representing the proportions of the major pelagic sediment components for at least the late Cenozoic.

Volcanogenic sediment occurs mostly around volcanic islands, although there is some transport from volcanos on the continents. *Miskell et al.* [1985] determined accumulation rates for volcanic indicators recorded from smear slides from DSDP cores, and *Howell and Murray* [1986] used this information to estimate the total volume of volcanogenic sediment in the ocean to be 2.9×10^6 km^3 (solid), which, using their value of 2.2 g cm^{-3} for the density, equals a mass of 6.4×10^{21} g. Inspection of the volcanic accumulation rates reported by *Miskell et al.* [1985] suggests that the variance within the DSDP data is so great that its interpretation as a global mass/age distribution would be misleading.

Bottom-transported terrigenous sediment could be calculated as a residual if the total sediment and the other four components were known. This method was used by *Howell and Murray* [1986], who determined the total volume of terrigenous material (both bottom-transported and "pelagic") on the ocean floor to be 70.3×10^6 km^3 (solid); they did not give a value for the density of this material, but, if it is assumed to be 2.7 g cc^{-1}, this equals 189.8×10^{21} g. Figure 1 includes our present state of knowledge of the mass/age distribution of the components of ocean floor sediments. The carbonate mass/age distribution is the only component which can be sketched out in any detail, but the absence of any appreciable quantity of pelagic carbonate older than 100 Ma is based less on widespread observation of sediments in the deep sea than on sediments exposed on land and the notion that the great flourishing of the calcareous plankton began in the Cenomanian [*Poldervaart*, 1955; *Hay and Southam*, 1975; *Sibley and Vogel*, 1976]. So little deep seafloor older than 100 Ma has been drilled that knowledge of the nature of early Cretaceous and Jurassic ocean sediments is still rudimentary. The opaline silica mass/age distribution is known only for the interval 15–85 Ma, using the fluxes determined by *Miskell et al.* [1985]. It should be noted that the total of 8.00×10^{21} g, obtained by adding the total fluxes for the ocean basins in Appendix 2 of Miskell et al. is considerably less than the 5×10^6 km^3 of biogenic silica on the ocean floor reported by *Howell and Murray* [1986] to have been calculated from the accumulation rate data of *Miskell et al.* [1985, Appendix 1]; 5×10^6 km^3 of opaline silica, using *Howell and Murray*'s [1986] value of 2.3 g cm^{-3} for the density, would have a mass of 11.5×10^{21} g. We note that *Miskell et al.* [1985, Appendix 2] report the fluxes in units of 10^{13} g cm^{-2} kyr^{-1} when the units must actually be 10^{13} g yr^{-1}. We are unsure of how *Howell and Murray* [1986] arrived at the sum of 5×10^6 km^3 of opaline silica on the ocean floor; compared with their estimation of 15.9×10^6 km^3 of calcium carbonate, this would mean that the ratio of opaline to calcium carbonate in deep-sea sediments would be 1:3, and experience indicates that this is untrue. The proportion for the late Cenozoic is closer to 1:10 [*Poldervaart*, 1955]. The flux data of *Miskell et al.* [1985] for opal during the interval 15–25 Ma, compared with the flux calculated for carbonate from *Sloan*'s [1985] data, have a ratio of about 1:10.

Howell and Murray [1986] reported the volume of calcium carbonate on the ocean floor to be 15.9×10^6 km^3, or 17%, of an ocean floor total sediment of 94.1×10^6 km^3. Using their assumed density of 2.72 g cm^{-3}, this equals 43.2×10^{21} g. The accumulation rate data of *Sloan* [1985] indicate that for the Cenozoic the calcium carbonate averages 42% by weight of the total sediment, and the accumulation rates for the Late Cretaceous given by *Southam and Hay* [1981] indicate that calcium carbonate is about 60% of the total Turonian-Maestrichtian sediment. Again, we are not sure of the source of this discrepancy.

Knowledge of the mass/age distribution of the major components of the ocean floor sediment is incomplete, because the information on sediments older than 85 m.y. is spotty and not necessarily representative of the total sediment of that age still in existence. Table 2 shows the available information on $CaCO_3$, calculated from the data of *Sloan* [1985] for the Cenozoic and from that of *Southam and Hay* [1981] for the Turonian-Maestrichtian. The estimates for opal accumulation for the interval 0–5 m.y. were derived from *DeMaster* [1981], assuming steady state for the input and output and assuming that the present flux of silica is the average for the past 5 m.y. The opaline silica accumulation rates shown for 5-m.y. intervals from 15–85 m.y. were derived from the flux data for 10-m.y. time increments from *Miskell et al.* [1985]. To divide the 10-m.y. data into 5-m.y. increments, we apportioned the total flux for the 10-m.y. interval into the 5-m.y. intervals, according to the area of the present ocean floor in existence at the midpoint of each of the 5-m.y. intervals.

Using this albeit incomplete knowledge of the mass of calcium carbonate present in sediments younger than 85 m.y. and the even more incomplete knowledge of the mass of opaline silica in sediments 0–5 and 15–85 m.y., it is possible to estimate the mass of pelagic sediment in two ways:

1. If we consider pelagic sediment to be typically 74% $CaCO_3$ [*Southam and Hay*, 1981], the total pelagic sediment will than be 1.35 times the mass of $CaCO_3$. The 5-m.y. interval masses from 0 to 85 m.y. are shown in Table 2. Ocean floor sediments of age 0–85 m.y. constitute 91.5% of the total ocean floor sediment mass. Dividing the estimate of pelagic sediment younger than 85 m.y. (129.6×10^{21} g) by 0.915, we can estimate the total mass of pelagic sediment to be 141.7×10^{21} g.

2. If we consider the red clay to be 19% of the total pelagic sediment [*Poldervaart*, 1955; *Southam and Hay*, 1977, 1981], then the total mass of pelagic sediment is 1.23 times the mass of calcium carbonate plus the mass of opaline silica. The ocean floor sediment of ages 0–5 and 15–85 m.y. is 74.9% of the total ocean floor sediment. The amount of pelagic sediment for these time intervals, estimated by the pelagic sediment equal to 1.23 (carbonate plus opal) formula, is 103.73×10^{21} g. Dividing this by 0.749 gives an estimate of 138.50×10^{21} g for the total of pelagic sediment.

These two methods yield remarkably close results. The average of these two values is 140.10×10^{21} g, so that 140×10^{21} g is a good approximation of the present state of knowledge.

DISCUSSION

The mass/age distribution is approximated by an exponential decay curve, so that the mass for any given 1-m.y. time increment is $(11.02 \times 10^{21}\ \text{g})e^{-0.0355t}$ Ma, where t is the midpoint of the 1-m.y. time increment. Expressed in terms of the 5-m.y. time increments in which the data are presented in Table 1, the equation is $(55.19 \times 10^{21}\ g)e^{-0.355t}$ Ma. This exponential decay shape of mass/age distribution of sediment on the ocean floor is a function of three factors: (1) existing area of ocean crust of a given age, (2) supply of sediment to the deep sea, and (3) submarine erosion and redeposition. Clearly, the area of ocean crust of a given age in existence at the present time dominates the mass/age distribution. This effect can be eliminated by considering only the changes in accumulation of sediment still in existence per unit area per unit time (given in columns 4 and 5 of Table 1; these numbers have been rounded and are thus smoothed with respect to the 5-m.y. interval accumulation values given in column 6). The changes in accumulation of sediment still in existence must

TABLE 2. Mass/Age Distribution of Carbonate, Opaline Silica, and Pelagic Sediment

	Carbonate			Opaline Silica	Estimates of the Mass of Pelagic Sediment		Red Clay	Terrigenous and Volcanic Sediment
Interval, Ma	Accumulation for Each 5-m.y. Interval, g cm^{-2}	Mass in Each 5-m.y. Interval, 10^{21} g	Cumulative Mass, 10^{21} g	Mass in 5-m.y. Intervals, 10^{21} g	1.35 × Carbonate Mass in Each 5-m.y. Interval, 10^{21} g	1.23 × (Carbonate + Opal) Mass in 5-m.y. Intervals, 10^{21} g	Pelagic Sediment Minus Carbonate Minus Opaline Silica Mass, 10^{21} g	Total Ocean Floor Sediment Minus Pelagic Sediment, Calculated as 1.23 × (Carbonate + Opal)*
0–5	4.38	12.00		2.791×10^{21}	16.20	18.19	3.46	31.22
5–10	3.68	9.44	21.44		12.74			(10.15)
10–15	3.51	8.49	29.93		11.46			(9.23)
15–20	3.39	7.73	37.66	0.577	10.44	10.22	1.91	8.03
20–25	3.55	7.63	45.29	0.545	10.30	10.05	1.88	6.27
25–30	4.10	8.36	53.65	0.248	11.29	10.59	1.98	3.41
30–35	2.60	5.00	58.65	0.234	6.75	6.44	1.21	3.46
35–40	2.29	4.16	62.81	0.261	5.62	5.44	1.02	6.81
40–45	3.25	5.58	68.39	0.247	7.53	7.17	1.34	7.45
45–50	3.25	5.26	73.65	0.286	7.10	6.82	1.27	7.18
50–55	2.82	4.28	77.93	0.268	5.78	5.59	1.04	6.43
55–60	2.88	4.08	82.01	0.064	5.51	5.10	0.96	5.11
60–65	2.45	3.21	85.22	0.060	4.33	4.02	0.75	2.41
65–70	4.71	5.69	90.91	0.055	7.04	7.06	1.32	0.60
70–75	2.36	2.61	93.52	0.051	3.52	3.27	0.61	1.31
75–80	1.45	1.44	94.96	0.034	1.94	1.81	0.34	1.25
80–85	1.74	1.56	95.52	0.030	2.11	1.96	0.37	1.12
Estimated totals			107.46†	7.94‡	141.69†	138.50‡	26.2	123.39
Proportion total oceanic sediment			41.0%	3.0%	54.1%	52.9%	10.0%	47.1%

*Except for 5–15 m.y., where terrigenous sediment is calculated as total ocean floor sediment minus 1.35 × carbonate.
†Total projected by assuming that the intervals for which data exist (0–85 m.y.) are 91.5% of the total.
‡Total projected by assuming that the intervals for which data exist (0–5, 15–85 m.y.) are 74.9% of the total.

represent the combination of the effects of changing the supply of sediment to the ocean floor and of submarine erosion and redeposition. The cumbersome term "accumulation of sediment still in existence" is used rather than "accumulation rate" in recognition of this complex combination of factors affecting the amount of sediment still in existence. There have undoubtedly been pulses of high sediment input, notably in the last 5 m.y., due to the rapid changes in sea level associated with the waxing and waning of continental ice caps, and in the interval from 85 to 90 m.y. ago, which includes Santonian, Coniacian, and Turonian sediments. The high accumulation still in existence representing the 85- to 90-m.y. interval may be in part a function of the time scale [*Hallam et al.*, 1985], which allots spans of only 1.5 m.y. each to the Coniacian and Turonian; alternatively, the very existence of these stages as recognized units may be related to the unusual conditions which produced much larger than average sediment thicknesses in a short time. Accumulation still in existence minima occurred within the Oligocene (30–35 m.y.) and Paleocene (60–65 m.y.). Times of low sediment supply to the ocean floor [*Davies and Worsley*, 1981] are also times of submarine erosion and hiatus formation [*Moore and Heath*, 1977; *Moore et al.*, 1978], so that it is not possible to separate these two factors using data on existing oceanic sediments alone. The analysis of DSDP data by *Sloan* [1985] indicates that times of high sediment accumulation rate are also times when terrigenous sediments dominate over pelagic sediment; during times of low overall accumulation rate, pelagic sediments dominate over the terrigenous.

Our estimate of the mass/age distribution of ocean floor sediment is determined by multiplying the amount of sediment still in existence by the relevant area of sea floor. A hiatus is a period of little or no sediment accumulation or subsequent removal and is therefore directly related to the amount of sediment still in existence representing a given time interval. The relationship between sediment in existence and hiatuses is apparent in Table 1 (column 4), where one can see that an increase in the amounts of sediment in existence representing the interval from 40 to 60 Ma closely corresponds to a broad decrease in the frequency of hiatuses occurring during this time interval, as shown by *Moore and Heath* [1977].

The overall composition of the sediments on the ocean floor can be estimated from proportions of pelagic and terrigenous sediment and is given in Table 3. For the terrigenous component we have used the composition for shelf slope and rise sediments given by *Sibley and Wilband* [1977], and for the pelagic component we have used the composition computed by *Southam and Hay* [1977], based on analyses by *Sujkowski* [1952]. Using portions of 52.9% pelagic and 47.1% terrigenous sediment respectively, the ocean floor sediments retain a distinctive composition, significantly different from that of any other major sedimentary body and from the composition of

TABLE 3. Composition of Ocean Floor Sediment, Its Components, and Average Sedimentary Rock

	Terrigenous Component*	Pelagic Component†	Ocean Floor Sediment‡	Average Sedimentary Rock§
SiO_2	53.3	24.1	37.8	59.7
Al_2O_3	11.6	7.4	9.4	14.6
Fe	3.5	4.4	4.0	4.5
MgO	3.2	1.3	2.2	2.6
CaO	9.6	40.2	25.8	4.8
Na_2O	1.0	1.4	1.2	0.9
K_2O	2.4	1.1	1.7	3.2
CO_2	8.6	16.8	12.9	4.7
Other	6.8	3.3	4.9	5.0

*From *Sibley and Wilband* [1977].
†From *Southam and Hay* [1977].
‡Calculated as 47.1% terrigenous and 52.9% pelagic.
§From *Garrels and MacKenzie* [1971].

the average sedimentary rocks of *Garrels and Mackenzie* [1971], also shown in Table 3.

As is apparent from the compilation of lithologies incorporated into Mesozoic-Cenozoic orogenic belts [*Spencer*, 1974], the obduction of pelagic sediment does not occur in any significant quantities, and most obducted ocean floor sediment has a terrigenous composition. However, it can be estimated that if the overall mass has remained constant through time, between 150 and 250 $\times 10^{21}$ g of ocean floor sediment are subducted every 180 m.y.. If, for simplicity, we assume 180 $\times$ $\times 10^{21}$ g are subducted every 180 m.y., this means that 1 $\times$ $\times 10^{21}$ g is subducted every million years. Some of this material is underplated on the roof of the subduction zone to become blueschists, another portion is transformed into andesitic volcanics, another portion is melted and incorporated with other rock to become intrusives, and another fraction may be mixed back into the mantle. Thus 1×10^{21} g per million years must be an approximate minimum for the loss of sediment from the sedimentary cycling system. This is essentially the same as the estimate of an upper limit of $1.1 \pm 0.5 \times$ $\times 10^{21}$ g per million years, cited as the quantity of sediments available for crust/mantle exchange by *Veizer and Jansen* [1985] from a different line of reasoning. It implies that if plate tectonics have been operating since the Archaean, a mass of sediment larger than the total mass presently in existence (about 2500×10^{21} g, according to *Southam and Hay* [1981]) has been subducted over the course of the Earth's history. Because much of the material being subducted is pelagic sediment and has a composition different from the overall composition, it would cause the entire sedimentary system to evolve in such a way as to become depleted in calcium and enriched in silica alumina and potash. However, the mass of sediment on the ocean floor at present may be anomalously large if, as suggested by *Poldervaart* [1955], *Hay and Southam* [1975] and *Sibley and Vogel* [1976], the calcareous plankton have only been contributing calcium carbonate to the deep sea for the past 100 m.y. If there were no calcium carbonate delivered to the deep sea by plankton and if all of the output were on continental shelves (assuming space to be available) the steady state value for the total mass of sediment on the ocean floor would be about 162×10^{21} g. If most of this were subducted rather than obducted as the plate tectonic machine operated, it still implies that a mass of sediment approximately equivalent to the mass of sediment presently in existence has been lost from the sedimentary cycling system to become metamorphic or igneous rock. In this scenario the material being subducted would have a composition similar to the main body of sedimentary material and would not be a driving force in the evolution of sediments. However, it may be that the factor exerting primary control on delivery of material to the ocean floor is the space available on subsiding passive margin shelves. Material which cannot be accommodated on the shelves must be passed on to the ocean floor. *Hay et al.* [1988] suggest that because calcium carbonate can be derived from ocean water along the entire length of the shelves and on carbonate banks whereas detrital sediment is injected at point sources (river mouths), the continental shelf and carbonate bank system may have operated more efficiently to remove carbonate prior to the appearance of abundant oceanic calcareous plankton 100 Ma.

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